13. THE PHYSICAL VOLCANOLOGY OF MARS

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Several types of constructive volcanic landforms (central volcanoes, tholii, the highland paterae, Alba Patera and small domes), as well as plains units of probable volcanic origin, exist in both the northern and southern hemispheres of Mars. Volcanism in the central highlands appears to have been explosive in character, while most of the constructive activity in the northern plains was effusive. In addition, highland volcanism appears to be relatively old compared to that in the northern hemisphere. Alba Patera appears to be unique in preserving both styles of volcanism and being transitional in age. Volcanism has had a significant influence on both the Martian cryosphere and atmosphere; numerous examples of volcanic-ice interactions exist within Elysium Planitia and to the south of Hadriaca Patera, and volcanism is believed to have contributed large volumes of water vapor and other gases to the evolving Martian atmosphere. Theoretical models can be used to predict the style of activity and the extent of the deposits produced by eruptions of various plausible magmatic-fluid combinations on Mars: if the magmas were rich in volatiles there would have been a greater likelihood of explosive activity taking place on Mars than on the Earth. There is evidence for the existence of large magma chambers and very high effusion rate eruptions on Mars, based on the volume of observed calderas and lava flows, and the occurrence of sinuous rilles. Tectonic deformation associated with volcanic constructs is primarily a consequence of loading and magma transport, while deformation in the volcanic plains reflects stresses associated with Tharsis and major impact basins. Despite the wealth of information that has been derived about Martian physical volcanology from Viking Orbiter data and related numerical modeling, several key questions remain unanswered. The lack of knowledge about the range of magma chemistries, the frequency and duration of individual eruptions, the longevity of activity at individual volcanoes, the reason for the lack of recent constructive volcanism in the southern highlands, and a lack of detailed knowledge of the meter- to decameter-scale topography and regional slopes of Martian volcanic landforms precludes a truly rigorous analysis of volcanic landforms on Mars. Some of these gaps in our current knowledge may be filled by continuing studies of high-resolution images and new data from the Mars Observer mission, but additional information from carefully selected sample-return sites will be needed to substantially improve our understanding of the physical volcanology of Mars.

In this review, the physical volcanology of Mars is considered to include the diversity of volcanic landforms, the implied styles of eruption associated with the construction of these landforms, the inferred internal structure of the volcanoes, and the influence that the eruptions have had on the Martian environment (both local and global in scale). The regional tectonics of volcanic landforms are considered in chapter 8, but the important deformational features locally associated with the volcanic constructs and plains are considered here because they provide insights into the internal structure of Martian volcanoes and the volumes and mechanical properties of the plains. Similarly, the regional stratigraphy of Mars, which contains abundant information on the rates of volcanic resurfacing through time, is described in chapter 11; the rate of production of volcanic materials is in turn related to the thermal history of the planet (see chapter 5). However, several of the specific geologic units that can be identified from this global mapping help to define the likely diversity of volcanic activity on Mars, as well as the rates at which magma and juvenile volatiles were released. These aspects of global volcanic stratigraphy are also considered here. Global spectroscopic data (chapter 17) together with measurements from the Viking Landers (Arvidson et al. 1989a) and geochemical models (McGetchin and Smyth 1978) support the concept originally proposed on morphological grounds that Mars has been dominated by mafic and ultramafic volcanism. The volcanic plains of the southern highlands are not given comprehensive treatment here, as few details are known of their eruptive processes and rates; Greeley and Spudis (1981) provide an excellent and still current review of Martian plains volcanism.

1. MORPHOLOGIC TYPES OF MARTIAN VOLCANIC FEATURES

Much of our knowledge of the physical volcanology of Mars comes from the 15 years of morphologic analysis of Mariner 9 and Viking Orbiter images (cf. Carr 1973, 1981). The morphologies of Martian volcanoes in general have strong analogies to those of volcanic landforms on Earth, and Martian
features are described using terrestrial terminology (see, e.g., Greeley and Spudis [1981], Cas and Wright [1987, pp. 27-30], and the glossary in this volume for definitions). In a particularly good early Viking-based review, Greeley and Spudis (1981) subdivided the main morphological types of volcanic features on Mars as listed in the following subsections.

A. Central Volcanoes

1. Shields. The Tharsis Montes (e.g., Arsia Mons) and Olympus Mons are the best known of this category. Early in the 1970s the Mariner 9 mission revealed that members of this class of volcano have many general similarities to basaltic shields in Hawaii (Carr 1973), possessing nested summit calderas and having numerous lobate lava flows on their flanks. Such Martian volcanoes differ from their terrestrial equivalents by virtue of their great size: Olympus Mons rises more than 25 km above the surrounding plains (Fig. 1) and has a basal diameter in excess of 600 km (Wu et al. 1984). Based on the overall topography of the volcano Elysium Mons, which apparently has shallow sloping outer flanks and a steeper summit region than Olympus Mons, Malin (1977) and Pike (1978) suggested that some Martian montes may be composite volcanoes rather than shields analogous to those found in Hawaii.

2. Domes (Tholi). These are smaller volcanoes than the montes. The tholi are generally dome shaped, and have relatively steep peripheral flanks (in some cases, flanks exceed 8° in slope). Typically the lower flanks of the tholi have been buried by flows of the surrounding plains, so that the true areal extent and vertical dimension of the volcano are not known. An analysis of the slopes and the sizes of the summit calderas of tholi in the Tharsis region suggest that these volcanoes may have been partially buried by up to about 4 km of lava (Whitford-Stark 1982). In some instances, such as Hecates Tholus (Fig. 2), these volcanoes appear to have experienced explosive activity (Mouginis-Mark et al. 1982b).

3. Highland Paterae. These are low relief shields with irregular summit craters and numerous radial channels on their flanks. West (1974) first attributed their topographic profiles to explosive volcanism, suggesting that they were made of ash flow deposits rather than lava flows. Subsequently, Greeley and Spudis (1981) and Greeley and Crown (1990) have carried out more detailed mapping of Tyrrhena Patera and, with the exception of a few lava flows close to the summit, agree that the flanks of this volcano are most likely composed of unconsolidated materials, probably ash deposits (Fig. 3). Earth-based radar topography of Tyrrhena Patera (Fig. 4) shows that the northern flanks have very shallow slopes (<0°25) over a distance in excess of 100 km, and that the summit of the volcano probably rises < 1 km above the surrounding plains.

4. Alba Patera. This volcanic landform appears to be unique on Mars (Carr 1973; Greeley and Spudis 1981). Although many of the flanks (Carr et al. 1977a) and the summit area (Cattermole 1987) consist of lava flows (Fig. 5), numerous digitate channels are found on the northern flanks that, together with thermal inertia data, suggest a pyroclastic origin for these materials (Fig. 6).
Fig. 3. Tyrrhena Patera is one of several highland volcanoes that appears to have formed predominantly by ash eruptions. Greeley and Spudis (1981) and Greeley and Crown (1990) interpret the summit area to be covered by lava flows, but prominent flow lobes and diagnostic surface morphologies are absent here, so that the summit may comprise welded-ash deposits. Prominent on the flanks of the volcano is a set of deeply eroded channels that emanate from the summit. Image width is equivalent to ~130 km. (Part of JPL Photomosaic 211–5730.)

Fig. 4. Earth-based radar topography of Tyrrhena Patera shows that the slopes on the northern flanks (top of figure) of the volcano are very low. Although the volcano has a basal diameter of ~600 km, the summit rises < 1 km above the surrounding plains. Note that the contours are interpolated between groundtracks, and assume an azimuthally symmetric volcanic cone. Contours are in km, solid horizontal lines mark positions of radar groundtracks. Larger adjacent impact craters (solid circles) are also shown. Data collected by Downs et al. (1975).
Fig. 5. The summit area of Alba Patera, centered at 40°6' N, 110°7' W. Numerous small lava flows (arrow) and wrinkle ridges (R) can be found on the near-summit flanks, supporting the idea of late-stage effusive volcanism associated with tectonic deformation. Image width is equivalent to 220 km. North is to the top. (Part of Viking Orbiter Photomosaics MTM-40107 and -40112).

Together with the low aspect ratio of the flanks (Alba Patera has a basal diameter > 1600 km, but rises only ~3 to 4 km above the surrounding plain), it appears likely that the unique nature of Alba Patera is due to its polygenic style of activity, with early explosive eruptions followed by subsequent effusive activity (Mouginis-Mark et al. 1988).

5. Numerous Small Domes. Many have been identified across Mars. Most enigmatic are the thousands of subkilometer-sized hills that exist within the northern plains (Frey et al. 1979; Frey and Jarosewich 1982). In addition to these lowlands plains hills, Plescia (1981) and Tanaka and Davis (1988) have identified several breached cones in Tempe Terra and Syria Planum that are morphologically similar to cinder cones produced by pyroclastic activity on Earth. Some of these appear to have lava flows emanating from them, strongly suggesting a volcanic origin for the cones.

B. Volcanic Plains

1. Simple Flows. These materials form areally extensive plains that generally contain wrinkle ridges similar in morphology to terrestrial and lunar mare ridges (Lucchitta and Klockenbrink 1981; Plescia and Golombek 1986; Sharpton and Head 1988; Watters 1991). There has been much debate about

Fig. 6. (a) Map of Alba Patera, showing the area from 32°5' N to 47°5' N and 100°7' W to 122°5' W, and the grabens that almost surround the summit. Outlined are areas shown in Figs. 5 and 6B. Also shown are the parts of the volcano where the digitate channels are found (stippled area), large meteorite craters (barbed circles), and areas where cloud cover or poor resolution precluded positive identification of certain surface features (dashed lines). (Mapped from U.S. Geological Survey Viking Orbiter photomosaics MTM-35102, -35107, -35112, -35117, -40102, -40107, -40112, -40117, -45102, -45107, -45112, -45117 and JPL Mosaic 211-5071.) (b) Map of channel networks on the northeast flank of Alba Patera. Outlined area C denotes the location of the stereo pair shown in Fig. 6c (figure from Mouginis-Mark et al. 1988). (c) High-resolution stereo pair showing a digitate channel network northeast of the summit caldera of Alba Patera (see Fig. 6B for location). Direction of flow is towards top of image. Note the lack of evidence of lava flows, which argues against these channels being of volcanic origin. Image resolution is ~8 m/pixel. (Viking Orbiter frames 445B07 (right) and 445B08.)
the origin of the ridged plains. A volcanic origin is preferred, based primarily on morphologic similarity to the lunar maria (Fig. 7). In a few places, sinuous rilles can be found in intimate association with the ridged plains, such as in southern Lunae Planum and Syrtis Major (Schaber 1982), which also suggest that these plains are volcanic. The term "simple flows" implies single cooling units (Walker 1972), although it is not clear whether these Martian plains consist of high-volume, single-cooling units (possibly flood lavas) of great extent (Greeley and Spudis 1981).

Recent depolarized Earth-based radar data for much of the equatorial belt of Mars (Thompson and Moore 1989a) reveal that the ridged plains have a much lower radar echo strength than the lava plains surrounding the shields. Such data may indicate that the ridged plains are comparable to terrestrial pahoehoe flows, while the lava plains may comprise topographically rougher aa flows (Gaddis et al. 1989). If this interpretation is correct, then radar cross sections may provide a useful indirect method for estimating the mass eruption rates of different volcanic materials due to the correlation between flow roughness and eruption conditions (Rowland and Walker 1990).

However, these radar data are ambiguous, because the ridged plains have radar cross sections comparable to most of the other (probably sedimentary) plains units on Mars (Table I).

Thicknesses of the ridged plains have been measured to constrain eruptive volumes. Thickness estimates for Lunae Planum, determined using either

![Fig. 7. Ridged plains in Hesperia Planum closely resemble the lunar maria, suggesting a volcanic origin for these Martian plains (Scott and Carr 1978). North is to the top left; image width is equivalent to 107 km. (Viking Orbiter image 418339, centered at 28°56'S, 240°34'W.)](image)

**TABLE I**

<table>
<thead>
<tr>
<th>Unit Depolarized Echo Strengths for Selected Martian Volcanoes, Lava Flow Fields and Plains Units.*</th>
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<tbody>
<tr>
<td><strong>Volcanic Constructs with Lava Flows on Flanks</strong></td>
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<tr>
<td>Ascalap Mons</td>
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<td>Arsia Mons</td>
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<td>Olympus Mons</td>
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<td>Alba Patera</td>
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<tr>
<td>Elysium Mons</td>
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<tr>
<td><strong>Volcanic Constructs with Channelized Flanks</strong></td>
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<td>Hecates Tholus</td>
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<td>Tyrrhena Patera</td>
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<td>Hadriaca Patera</td>
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<tr>
<td><strong>Lava Flows on Volcano Flanks</strong></td>
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<tr>
<td>Olympus Mons Base</td>
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<tr>
<td>Elysium flows</td>
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<tr>
<td>Alba Northwest</td>
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<tr>
<td>Alba Southwest</td>
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<tr>
<td>Distal Alba flows</td>
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<tr>
<td><strong>Ridged Plains</strong></td>
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<tr>
<td>Syria Plassum</td>
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<tr>
<td>Syrtis Major</td>
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<tr>
<td>Hesperia Planum</td>
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<tr>
<td><strong>Plains Materials of Questionable Origin</strong></td>
</tr>
<tr>
<td>Arcadia Planitia</td>
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<tr>
<td>Acidalia Planitia</td>
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<tr>
<td>Isidis Planitia</td>
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*Note that volcanic constructs with lava flows (e.g. Olympus Mons) have echo strengths more than an order of magnitude larger than the highland patera, and that most simple flow units have much lower returns than complex flow materials (table from Thompson and Moore 1989a).

1Echo strength is variable $A$, defined by Thompson and Moore, and is given in equivalent total radar cross section.
the exposed rim heights of partially buried impact craters or the smallest diameter crater that survived a resurfacing event, range from about 250 m to 1.5 km (De Hon 1981, 1982, 1988; Saunders and Gregory 1980; Saunders et al. 1981; Frey et al. 1988). Independent estimates of the thicknesses of these units have also been made where they are exposed on the walls of the Valles Marineris and Kasei Valles. Values range from < 1 km to approximately 2 km (Nedell et al. 1987; Robinson and Tanaka 1988; Watters 1991; chapter 14).

2. Complex Flows. These materials have multiple overlapping flow lobes, and are usually found around the periphery of the shield volcanoes (Fig. 8). Greeley and Spudis (1981) suggested that this type of plains material resulted from the eruption of more sporadic, lower effusion-rate eruptions compared to the simple flows. Detailed geologic maps of flows of this type within the Tharsis region have been prepared by Scott and Tanaka (1986), and the identification of many flows originating several hundreds of km from the large volcanic constructs (Schaber et al. 1978; Mouginis-Mark et al. 1982a) suggests that vents, fissures and feeder dikes are quite numerous within much of Tharsis. The complex flows can extend for >1000 km from the summit calderas of the shields, but have smaller areal extent than the simple flow plains. High-resolution (10 to 20 m/pixel) images of these flows show numerous festoon ridges and lava channels (Theilig and Greeley 1986), and Earth-based radar-scattering data indicate that these flows are very rough at a radar wavelength of 12 cm (Schaber 1980).

3. Undifferentiated Flows. There are several instances where plains materials are located close to a volcanic construct, but the materials are of uncertain origin. Greeley and Spudis (1981) discuss the plains units around the margins of Tharsis and Elysium, and note that although there is no unequivocal evidence for a volcanic origin for these plains, there is also no compelling evidence for an origin by other processes. The northern plains of Mars constitute most of this class, and in several places, such as Ismenius Lacus (Lucchitta 1978c), wrinkle ridges and occasional flow lobes are seen.

4. Plains of Questionable Origin. Particularly within the southern highlands, there are many eroded-plains materials (approximately 42% of the total surface area of the cratered hemisphere of Mars; Greeley and Spudis 1978) that may have a volcanic origin but are too heavily eroded to permit the confident identification of flow fronts and other diagnostic features. The heavily fractured plains around Tharsis is a second example of questionable materials that may be of volcanic origin.

II. STRUCTURE OF MARTIAN VOLCANOES

Considerable information regarding the internal structure of terrestrial volcanoes has been gleaned by analysis of seismic data (Ryan 1988), by the detailed mapping of vent distributions (Porter 1972; Nakamura 1977; Munro and Mouginis-Mark 1990), from analysis of exposed dike swarms (Walker 1987; Knight and Walker 1988), from laboratory models of volcanic edifices (Fiske and Jackson 1972) and from numerical modeling of eruptive episodes (Wilson and Head 1988a; Head and Wilson 1989). For Martian volcanoes, the dominant observation is that the giant montes have relatively simple morphological structures compared with their terrestrial counterparts, apparently being mainly controlled by deep-seated regional tectonic forces. This is in contrast to terrestrial volcanic centers (e.g., Hawaii, Galapagos Islands, Reunion Island), where multiple volcanoes may be active in close proximity to each other, with the resultant formation of rift zones due to the interference of local stress fields (Fiske and Jackson 1972), and the production of flanks liable to collapse due to oversteepening (Duffield et al. 1982). Maps of the distribution of vents on Arsia, Pavonis and Ascraeus Montes (Carr et al. 1977a; Crumpler and Aubele 1978) have nevertheless helped reveal incipient

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Fig. 8. A sequence of lava flows to the west of Hecates Tholus in Elysium Planitia. The direction of flow is towards the top of this image. Image resolution is 52 m/pixel. The width of the image is equivalent to 115 km. North is towards the top of the image. (Mosaic of Viking Orbiter images 651A07-12; centered at 33°N, 215°W.)
rift zones trending roughly perpendicular to the major NE-SW trend of the Tharsis ridge volcanoes.

The summit areas of the Martian shields also reveal information on the internal structure of the volcanoes (Fig. 9). The large (>60-km diameter, 2- to 3-km deep) nested summit calderas of Olympus Mons and Ascrœus Mons show that multiple collapse events were associated with each volcano, presumably due to large-volume flank eruptions partially evacuating each magma chamber (Mouginis-Mark 1981). The topography of the summit area of Olympus Mons (Fig. 9b) reveals that the rim of the youngest collapse caldera is also topographically higher (by >2 km) than the other parts of the caldera rim. This correlation between rim elevation and age of collapse, and the absence of young lava flows infilling older adjacent segments of the caldera, suggest that dike intrusions have played a major role in volcano growth in a similar manner to certain terrestrial shields (Walker 1987, 1988; Mouginis-Mark and Mathews 1987). Only small-volume lava flows can be identified at the summits of Olympus Mons and Ascrœus Mons (Mouginis-Mark 1981; Zimbelman 1985), but one of the more enigmatic aspects of Martian volcanism is the implied size of the parent magma chambers. Due to the very large volume of many of the lava flows associated with the Tharsis volcanoes (>100 km³; Wood 1984a; Cattermole 1987), magma chambers of Martian volcanoes are believed to be up to 3 orders of magnitude larger in volume than their terrestrial counterparts. The spacing of volcanic centers in Tharsis suggested to Whitford-Stark (1982) that the size of each volcano directly correlates to the distance from its nearest neighbors, and Whitford-Stark proposed that in the case of the groups of smaller Tharsis volcanoes (Biblis/Ulysses Patera, Ceraunius Tholus/Uranius Tholus/Uranus Patera), these volcanoes shared common magma sources that inhibited the large-scale growth of any of the individual volcanoes. The relatively large diameters of the summit calderas of Biblis, Ulysses and Uranus Paterae nevertheless suggest that the size of the magma chamber within these volcanoes was probably as large as those within other Martian volcanoes such as Pavonis Mons (Wood 1984b).

To the west of Elysium Mons there is a complex area of hummocky terrain which includes lava flows, vents, fissures, collapse pits and sinuous rilles (Mouginis-Mark et al. 1984). This area appears to be a likely candidate for a failed volcanic construct where, perhaps because of the high regional slope or the high mass eruption rate of the magmas, the volcano was unable to produce the more common shield topography.

III. ERUPTIONS THROUGH TIME

A. Age of Volcanoes

Several different cratering curves have been developed that place the ages of Martian volcanic landforms in relative chronologies (see Tanaka et
al. 1988; chapter 12). Tanaka et al. (1988) developed a resurfacing chronology for Mars, and inferred that volcanic surfaces (\( \sim 84 \times 10^9 \) km\(^2\)) cover more than half of Mars. Pescia and Saunders (1979) and Neukum and Hiller (1981) have also carried out crater counts for the flanks of the Martian volcanoes, and both identified Olympus Mons as one of the youngest features and the highland paterae as being generally old. Of particular interest is the range of intermediate ages of parts of the flanks of Alba Patera, indicating that this volcano had a protracted history that spanned the period of early southern-hemisphere volcanism and late-stage shield building (Neukum and Hiller 1981). Extremes in the age of volcanic events include the speculative identification of highland volcanoes in the western hemisphere of Mars (Scott 1982), and the possible very recent activity on Hecates Tholus (Mouginis-Mark et al. 1982b) and in the floor of Valles Marineris (Lucchitta 1987a, b).

**B. Mechanism of Volcanic Resurfacing**

The origin of the major young volcanic centers on Mars remains enigmatic. Other chapters address in detail the geophysical (chapter 5) and petrologic (chapter 6) constraints that can be placed on the early volcanic activity on Mars. Here, we note two key observations: the location of the most recent volcanism on Mars in the northern lowlands, and the absence of lava shieldbuilding activity in the southern highlands. In addition, any mechanism proposed to account for the extensive resurfacing in the north should also explain the approximately 3-km elevation difference between the northern and southern hemispheres. If the elevation change is isostatic, as suggested by gravity models (Phillips 1988), then it is appealing to invoke an endogenic mechanism such as subcrustal erosion, occurring perhaps in response to vigorous mantle convection (see, e.g., Wise et al. 1979; Turcotte 1988). Such a mechanism would "thin" the crust above a broad thermal anomaly and would be consistent with a concentration of volcanism in the lowlands. Alternatively, it has been proposed that a giant impact (Wilhelms and Squyres 1984) or multiple impacts (Frey and Schultz 1988) could have created thermal and mechanical weakening in the deep lithosphere that could have promoted continued lowlands volcanism.

**IV. INFLUENCE OF VOLCANISM ON THE MARTIAN ENVIRONMENT**

**A. Input to the Atmosphere**

Volcanoes on the Earth typically inject large quantities of volatiles (water, carbon dioxide, sulfur dioxide and lesser amounts of other compounds including halogens) into the atmosphere, and it is likely that Martian volcanoes had a similar influence. Slow release of sulfur compounds and carbon dioxide also commonly takes place from shallow magma reservoirs through overlying fracture systems during repose periods between eruptive episodes.

Settle (1979) considered the formation and deposition of juvenile volcanic sulfate aerosols on Mars, and concluded that volcanic so2 released at the surface would result in sulfate aerosol formation, and that the upper atmosphere circulation would cause these particles to be distributed on a global or hemispheric scale. Such a process could account for the high concentrations of sulfur (\( 3 \) to 4 \%w) within the surficial soils studied at the two Viking Lander sites. The release of water from effusive activity on Mars as a function of geologic time was addressed by Greeley (1987), who estimated that for the preserved geologic record a total amount of water equivalent to a layer \( \sim 50\)-m thick could have been degassed from the plains and constructional volcanic materials preserved on the Martian surface. Based on the exposed surface area of these materials, Greeley (1987) estimated that the greatest volume of juvenile water (equivalent to a layer \( 16.3\)-m thick distributed planet-wide) was probably released during Early Hesperian times, with significant additional amounts during Late Hesperian (\( \sim 11.1\) m) and Early Amazonian (\( \sim 7.7\) m) (Table II).

The effect of volcanic outgassing of water vapor and sulfur dioxide on the early Martian atmosphere has also been considered by Postawko and Kuhn (1986). Although it is likely that sulfur dioxide has been released into the Martian atmosphere at various times from volcanic eruptions (Settle 1979), the actual amount is uncertain and, due to the low atmospheric density, the lifetime of sulfur dioxide in the atmosphere may have been as short as \( \sim 10\) yr. Because it is a net absorber of solar radiation, the effect of this sulfur dioxide in the atmosphere would most likely have been to act as a trigger to increase the amount of water vapor in the atmosphere by greenhouse warming (Postawko and Kuhn 1986).

**TABLE II**

<table>
<thead>
<tr>
<th>Age</th>
<th>Volcanic Materials ( (10^9 ) km(^2))</th>
<th>Water Layer* (m)</th>
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<tr>
<td>Late Amazonian</td>
<td>5.3</td>
<td>1.2</td>
</tr>
<tr>
<td>Middle Amazonian</td>
<td>20.78</td>
<td>4.8</td>
</tr>
<tr>
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<td>7.7</td>
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<tr>
<td>Late Hesperian</td>
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</tr>
<tr>
<td>Early Hesperian</td>
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</tr>
<tr>
<td>Late Noachian</td>
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</tr>
<tr>
<td>Middle Noachian</td>
<td>0.28</td>
<td>0.1</td>
</tr>
<tr>
<td>Totals</td>
<td>201.69</td>
<td>46.0</td>
</tr>
</tbody>
</table>

*Table after Greeley 1987.

*Values are given for the thickness of a water layer that would completely cover the planet. A layer of water 1-m thick is equivalent to a water volume of \( 0.144 \times 10^9 \) km\(^3\).
Specific amounts of released volatiles from individual volcanoes have also been estimated by Wilson and Mouginis-Mark (1987) for Alba Patera. The morphology of the volcano flanks suggests that, early in the eruptive history of Alba Patera, explosive volcanism characterized this volcano (Mouginis-Mark et al. 1988). The explosive eruptions were probably driven by volatiles that were both juvenile and terrigenous in origin. If the main explosive activity was due to juvenile volatiles, then all of the eruptions of Alba Patera would have injected a volume of water into the Martian atmosphere that was equivalent to a layer –5 mm to 5 cm deep distributed over the surface of the entire planet (Wilson and Mouginis-Mark 1987). This volume is small compared to the 46 m calculated by Greeley (1987) for all water released by volcanism on Mars, but indicates the contribution of individual volcanoes to the global volatile inventory. Wind-directed fume from the summit of Alba Patera has been proposed as a possible mechanism for the asymmetric distribution of the channel networks on the northern flanks of the volcano (Mouginis-Mark et al. 1988).

B. Volcano and Near-Surface Volatile Interactions

The profusion of large volcanic plains and shields, together with the morphologic (Carr 1986) and thermal evidence (Kieffer et al. 1976b) for water and ice on Mars, strongly suggest that magma and water or ice would have interacted during the evolution of the Martian landscape. Allen (1979) and Hodges and Moore (1979) each suggested that certain features on Mars resemble landforms found in Iceland that are due to the interaction of basaltic magma and ice, and the basal materials of Olympus Mons have variously been described as subglacial in origin (Hodges and Moore 1979), or the result of early pyroclastic volcanism (King and Riehle 1974), perhaps akin to the early explosive volcanism associated with Alba Patera (Mouginis-Mark et al. 1988). Greeley and Spudis (1981) and Greeley and Crown (1990) have also proposed that phreatomagmatic eruptions were responsible for much of the construction of Tyrrhenia Patera.

More direct evidence for the interaction between magma and volatiles within the Martian regolith comes from the large numbers of probable fluvial channels that exist to the west of Elysium Mons (Mouginis-Mark 1985) and to the south of Hadriaca Patera (Squyres et al. 1987). In the case of the Elysium channels (Fig. 10), melt-water release appears to have been responsible for the generation of the large volumes of sediment that now partially infill the northern plains in Utopia Planitia (Lucchitta et al. 1986). Intrusive activity probably played a key role in the release of water adjacent to Hadriaca Patera (Squyres et al. 1987), and there are additional small (10 to 50 km long, <2 km wide) channels immediately to the east of Olympus Mons that could also be the result of water released by intrusive activity (Mouginis-Mark 1990).
V. THEORETICAL MODELING OF ERUPTION PROCESSES

A. General Considerations

The range of styles of volcanic activity on any planet are a complex function of the chemistry, and hence rheology, of the magmas produced at depth by partial melting, and of the environmental conditions acting on magmas as they approach the surface—mainly the acceleration due to gravity and the atmospheric pressure (Wilson and Head 1983; Wilson 1984). The relatively low gravity on Mars compared with the Earth's gravity reduces the buoyancy forces acting on partial-melt bodies, and so reduces the ascent speed of a melt body of a given size. This reduction in the ascent speed requires the formation of systematically larger diapiric bodies ascending through the deep mantle if such bodies are to avoid excessive cooling. The consequence should be the accumulation of systematically larger magma reservoirs in the shallow lithosphere. The low gravity should also lead to the formation of systematically wider dikes connecting the subsurface reservoirs to the surface (Wilson and Head 1988b; Wilson and Parfitt 1989), encouraging the more rapid effusion of magma of a given rheology.

Another consequence of the lower Martian gravity is that any specific lithostatic pressure is reached at a greater depth on Mars than on Earth. Since magma volatile solubility is mainly pressure dependent, gas-bubble nucleation and gas exsolution will begin at greater depths than on the Earth, encouraging efficient volatile release, especially of low solubility species like CO₂ (Table III). The low atmospheric pressure will lead to thorough degassing of Martian magmas as they approach the surface and lead to enhanced gas loss from lavas and pyroclastics immediately after their eruption. Any magma containing >5 to 10 ppm by weight of a relatively volatile substance (such as H₂O) is likely to be completely disrupted in a fire fountain, and will form pyroclastic particles entrained in their own released gas (Wilson and Head 1981a). As a result, given similar volatile contents, a larger fraction of all volcanic eruptions is more likely to be explosive on Mars than on the Earth, although this does not preclude the formation of extensive lava flows since coalescence of hot pyroclasts around vents can produce low-viscosity lavas (Head and Wilson 1989).

B. Explosive Activity

In view of the morphologic (Carr et al. 1977a; Carr 1981), spectroscopic (chapter 17), and Viking Lander (Arvidson et al. 1980a) evidence suggesting that Martian magmas were commonly mafic, it is important to explore the sizes expected for the common products of explosive activity in such magmas. Strombolian activity (defined as the intermittent explosive discharge of material due to coalescence of gas bubbles in magmas rising at relatively slow speeds in narrow vents [Wilson and Head 1981a]), will lead to a range of gas pressures and pyroclast speeds on Mars similar to the range for the Earth (Wilson et al. 1982). The dispersal of these clasts in the Martian atmosphere and gravity field should produce cinder cones with widths up to 200 m. The more nearly continuous accumulation of pyroclasts from Hawaiian-style fire fountains may produce ash and scoria cones with diameters in the range 200 m to 10 km, and some of the low cone-shaped to shield-shaped features found in Tempe Terra (Plescia 1981) and Syria Planum (Tanaka and Davis 1988) may represent the products of such activity.

On Earth, the majority of large-scale explosive eruptions driven by magmatic volatiles, especially of the plinian and ignimbrite-forming types, involve andesitic or silicic magmas (Walker 1973a). Francis and Wood (1982) have reviewed the morphologic and geochemical evidence for silicic volcanism on Mars, and concluded that there is no compelling evidence for such activity. However, the factors mentioned above that lead to more vigorous gas release and higher eruption rates in all magmas on Mars compared with those on Earth imply that it may have been common on Mars for basaltic magmas to produce plinian eruptions (the 1886 eruption of Tarawera, New Zealand, is a terrestrial example of such activity; Walker et al. 1984), or to generate ignimbrites. Additionally, the whole spectrum of smaller-scale explosive activity styles encountered on Earth in mafic magmas (Hawaiian, Strombolian, vulcanian, plinian) is also expected on Mars (Wilson and Head 1981b, 1983; Wilson et al. 1982).

Strong morphologic evidence for ash-fall deposits from plinian activity has been identified on Hecates Tholus (Mouglnis-Mark et al. 1982b). The rise height of a plinian eruption cloud is expected theoretically to be a function of the heat-injection rate at its base (which is in turn proportional to the mass-eruption rate of magma) and the pressure and temperature distribution.

### TABLE III

Calculated Gas Nucleation Depths and Magma Fragmentation Depths on Mars

<table>
<thead>
<tr>
<th>Volatile Content (wt. %)</th>
<th>Total</th>
<th>Nucleation Depth</th>
<th>Fragmentation Depth</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>H₂O in Basalt (m)</td>
<td>H₂O in Rhyolite (m)</td>
<td>CO₂ in Rhyolite (km)</td>
</tr>
<tr>
<td>5</td>
<td>21,000</td>
<td>13,000</td>
<td>196</td>
</tr>
<tr>
<td>3</td>
<td>10,000</td>
<td>4,800</td>
<td>118</td>
</tr>
<tr>
<td>1</td>
<td>2,100</td>
<td>530</td>
<td>39</td>
</tr>
<tr>
<td>0.3</td>
<td>380</td>
<td>48</td>
<td>12</td>
</tr>
<tr>
<td>0.1</td>
<td>80</td>
<td>5</td>
<td>4</td>
</tr>
<tr>
<td>0.05</td>
<td>14</td>
<td>0.4</td>
<td>1.2</td>
</tr>
</tbody>
</table>

*Values are given as a function of total volatile content of magma for three magma/volatile combinations. The solubility of CO₂ is essentially the same in basalts and rhyolites (table from Wilson et al. 1982).*
in the planetary atmosphere (Morton et al. 1956). This relation has been verified using observational data for terrestrial plinian eruption clouds (Settle 1978; Wilson et al. 1978). When the mean atmospheric pressure and temperature profiles of the Earth and Mars are inserted into the theoretical relation it is found (Wilson et al. 1982) that plumes fed by eruptions with a given discharge rate of magma from the vent are expected to rise to heights about 5 times greater than those on Mars. (Fig. 11). Models of the dispersal of pyroclastic particles falling from the plumes (Mouginis-Mark et al. 1988) show that mm-sized particles will commonly be carried 100 km downwind from the vent, and that 50 μm diameter particles may travel up to 1000 km.

Reimers and Komar (1979) recognized that several Martian volcanoes (such as Ceranius Tholus, Uranus Tholus, Uranus Patera and Hecates Tholus) have channel systems on their flanks, which they interpreted to be due to massive density currents, possibly pyroclastic flows, scouring the flanks of these volcanoes. Although many channels of this type are now believed to be of fluvial origin (Mouginis-Mark et al. 1982, 1988; Gulick and Baker 1987a), other examples of channels on the partly eroded flanks of Alba Patera suggest that flank deposits of this type may have been formed by pyroclastic volcanism early in the volcano's history (Mouginis-Mark et al. 1988).

More speculative debate concerns possible ignimbrites in Amazonis Planitia. These deposits remain controversial because their characterization as ignimbrites relies on their high albedo, easily eroded nature, large volume (~3.85 × 10^9 km^3), and their morphologic resemblance to sequences of welded and nonwelded terrestrial ash flows (Scott and Tanaka 1982). Other interpretations have been suggested (Schultz 1985), but the key problems in describing them as ignimbrites lie in the implied silicic magma chemistry (Francis and Wood 1982), the poorly defined source areas, and their very large surface areas and, hence, implied great travel distances. Run-out distances in excess of 2500 km would be needed for some of the older flows identified by Scott and Tanaka (1982) if they are indeed single ignimbrite units, whereas theoretical considerations (Wilson et al. 1982; Mouginis-Mark et al. 1988) suggest that maximum runout distances of 400 km for ignimbrite flows are more likely on Mars.

Finally, a variety of volcanic-explosion styles on Earth involve the interaction between magma and near-surface volatiles such as liquid water and ice. Both water and ice may be involved in such activity on Mars. When intruding magma vaporizes volatile materials trapped within an edifice, or lava advances over volatile-rich ground, the peak pressures reached are controlled by the strengths of the surrounding country rock layers or the cooling upper surface of the magma. Since these strengths are likely to be similar in all silicate rocks, it is expected that the initial velocities of explosion products will be essentially the same on Mars as on the Earth (Wilson 1980). The maximum travel distances of the products will be influenced by gravity and atmospheric drag, however, and discrete explosions on Mars should produce ballistic debris distributed over ranges up to 10 km. The fact that fallout of small clasts from any associated eruption clouds will also occur over a greater area (since the clouds rise higher on Mars) leads to the expectation that the landforms produced by such events will be wider and have less relief than on the Earth, making them harder to recognize in spacecraft images.

C. Lava-Producing Activity

Recognition of the existence of very long lava flows on Mars (see, e.g., Carr et al. 1977a; Pieri et al. 1985) has been an important spur in attempts to relate lava-flow length, width, planimetric area and thickness to lava rheology and effusion rate (Hulme 1976; Moore et al. 1978; Zimbelman 1985; Baloga and Pieri 1986; Pieri and Baloga 1986). These relationships are not well understood, even empirically, for the Earth (cf. Walker 1973b; Malin 1980; Pinkerton and Wilson 1988) and there is a strong incentive to develop a completely general model of lava-flow motion for application to all of the terrestrial planets (Wilson and Head 1983). Existing models suggest that the lengths of the longest flows are dictated mainly by magma eruption rates and
that, for a wide range of rheologies, it may not be safe to use the inferred
test to make a further inference about magma chemistry. The rate of
magma effusion feeding flows on Mars was commonly very high by terrestrial
standards ($>10^6$ m$^3$s$^{-1}$; Baloga and Pieri 1985; Cattermole 1987), being
comparable to that of terrestrial flood basalt or lunar mare eruptions. Most theo-
retical models can be made to fit ultra-basic to andesitic compositions (Zim-
belman 1985), provided that the volatile contents of Martian magmas were
similar to those of terrestrial magmas with the same major element composi-

Though many Martian flows were probably fed from long fissure-type
vents (see, e.g., Cattermole 1986), local concentration of eruptive activity
into shorter lengths of fissures would have provided the conditions for
more localized fire-fountains to form. In such eruptions, the combinations of
particle-size distribution, ejection velocity and total particle flux would lead to
optically dense conditions, pyroclasts being so closely crowded that
little radiative cooling to the surroundings could have occurred (Wilson and
Head 1981a, Head and Wilson 1989). These pyroclasts would have coalesced
on landing near the vents into lava ponds at near-magmatic temperatures
which in turn would have overflowed to feed lava flows. Analysis of the lava
motion in such ponds and flows (Wilson and Head 1981a; Wilson et al. 1982)
shows that turbulent conditions would prevail at the higher end of the inferred
range of effusion rates for Newtonian magmas (Fig. 12). The consequent

efficient transfer of heat to the underlying surface (Hulme 1973) would max-
imize the chance of thermal erosion taking place (Hulme and Fielder 1977)
to create sinuous-rille channels and their characteristic circular or elongate
source craters (Head and Wilson 1981). A total of 81 Martian sinuous rilles
have been identified planet-wide (Wilson and Mougins-Mark 1984), of
which 57 are located in Elysium Planitia and several other examples are
concentrated in Syrtis Major (Schaber 1982). The Elysium Planitia rilles (Fig.
13) range in length from 20 to 200 km, and for the larger ($>120$ km in length)
examples, the inferred mass eruption rates are in the range $1.2$ to $23.0 \times 10^8$
kg s$^{-1}$ (Wilson and Mougins-Mark 1984).

![Fig. 12. Eruption conditions in terrestrial basaltic scoria (left) and Martian (right) fire-fountain eruptions that result in thermal erosion of bedrock to form sinuous rilles. The solid curves show, as a function of mass eruption rate $M$ through the vent and exsolved magma content $n$, the radius in meters of the zone within which magma could would land hot enough to coalesce into a lava flow. The dashed lines are contours of equal Reynolds number for the motion of the lava draining from the resulting lava pond. If the Reynolds number exceeds about 1000, turbulent motion and rille formation will occur (figure from Wilson et al. 1982).](image1)

![Fig. 13. Distribution of lava flows (arrows) and sinuous rilles (solid circles show locations of source craters) within Elysium Planitia. Note that rilles occur on the flanks of Elysium Mons and the adjacent plains. Such a distribution probably indicates more than one near-surface magma chamber in order to accommodate the high-mass eruption rates believed to be associated with these eruptions (figure from Wilson and Mougins-Mark 1984).](image2)
VI. TECTONICS ASSOCIATED WITH MARTIAN VOLCANISM

A. Volcanic Constructs

Concentric graben that surround major volcanic constructs are interpreted to have formed due to loading of the Martian lithosphere (Comer et al. 1985). Knowledge of the mass of the load and the radial distance of the graben from the volcano center provides an indication of the thickness of the elastic lithosphere at the time of graben formation. Comer et al. (1985) and Hall et al. (1986) have analyzed the major Martian volcanic loads and found marked spatial heterogeneities in elastic thickness of the elastic lithosphere of Mars. These lithosphere thickness variations have implications for Martian internal structure and thermal evolution which are discussed in chapter 8.

Two of the most enigmatic features on Mars are the Olympus Mons aureole and basal escarpment (Fig. 14). The aureole consists of lobes of ridged materials distributed primarily to the north and northwest of the volcano that extend for distances of up to 1000 km. This feature has been interpreted as the erosional remnant of a pre-existing volcanic structure (Carr et al. 1972), and as a series of subglacial lava flows (Hodges and Moore 1979) or pyroclastic flows (Morris 1982). Other workers have proposed the aureole to have formed due to gravity sliding of material away from the basal escarpment (Harris 1977; Lopes et al. 1980, 1982; Francis and Wadge 1983; Head and Carras 1985; Tanaka 1985). The nature of the escarpment, which has a height of up to six kilometers above the surroundings, is also a matter of debate. Interpretations include both erosional (King and Riehle 1974; Head et al. 1976; Lopes et al. 1980; Tanaka 1981; Morris 1982) and tectonic (Harris 1977; Morris 1981b; Borgia et al. 1990) mechanisms.

The summit caldera complex of Olympus Mons represents one of the clearest examples of the relationship between tectonic and volcanic processes (Fig. 9). The central portion of the caldera, which is topographically low (Wu et al. 1984), is marked by numerous ridges, the largest of which are morphologically similar to wrinkle ridges found in the ridged plains and Lava plains. Inside the caldera rim, at a topographically high level, are circumferential graben (Mouginis-Mark et al. 1990; Watters and Chadwick 1990). The relationship of the summit topography to the tectonic features, in combination with photogeologic evidence for basalt-like resurfacing of the caldera floor (Greeley and Spudis 1981), is believed to indicate that a large lava lake cooled and subsided as magma in the underlying chamber was withdrawn by flank eruptions (Mouginis-Mark 1981). The state of stress implied by certain tectonic features within the caldera, which presumably formed due to floor subsidence associated with magma withdrawal, has been used to estimate the depth to the top of the magma chamber (Zuber and Mouginis-Mark 1990). The best-fit depth range (<16 km) constrains the magma chamber to have been within the volcanic edifice at the time of the subsidence episode. This

depth is much shallower than the predicted source depth of Martian magmas (Bertka and Holloway 1989). However, it is consistent with a result suggested from gravity scaling of the depths of magma chambers of terrestrial shields, assuming that chamber depth is controlled by magma buoyancy (Wilson and Head 1990; Ryan 1987).

The caldera complex at the summit of Ascreaus Mons consists of eight separate craters in various stages of preservation. The floor of the largest (and
youngest) crater contains numerous ridges that may be tectonic in nature, while the caldera rim exhibits mare-type wrinkle ridges and circumferential graben that may be related to detumescence and slumping of the caldera walls (Mougins-Mark 1981). The summit caldera of Arsia Mons is infilled by lava flows and there are no graben or ridges on the caldera floor; however, circumferential graben occur away from the rim (Crumpler and Aubele 1978). The calderas of Hadriaca Patera, Amphitrites Patera and Syrinx Major show similar structures. The two-level summit caldera of Pavonis Mons exhibits radial ridges within the caldera and concentric graben outside the rim (Crumpler and Aubele 1978) that may reflect subsidence of the summit. Details of the stress regimes and the possible implications for near-summit volcanic processes have not been quantitatively addressed for any of these structures.

On the flanks of Olympus, Ascreus, Arsia and Pavonis Montes are a series of terraces that are distributed concentrically about the summit calderas. Morphological analysis and models suggest that these features may be thrust faults caused by self-compression of volcanic edifice (Thomas et al. 1990). Subsequent models have shown that flexure of the lithosphere beneath a volcano due to the rapid emplacement of the volcanic construct would modify the magnitudes and principal directions of edifice stresses with time (McGovern and Solomon 1990). These models predict stress states consistent with summit eruptions shortly after load emplacement and flank eruptions at later times. Yet to be quantitatively addressed are the effects of volcano growth and associated lithospheric flexure for more realistic loading time scales, in which the state of stress of the volcano and surroundings changes during the shield-building process.

B. Volcanic Plains

Tectonic processes associated with probable volcanic flows are manifest in the ridged plains (Fig. 7). Wrinkle ridges in the plains can be explained by "global-scale" compressional stresses related to the development of Tharsis (Phillips and Lambeck 1980; Banerdt et al. 1982; Sleep and Phillips 1985; Watters and Maxwell 1986) and regional scale stresses associated with impact basins (Chicarro et al. 1985). While ridges are found in other geologic units, they are most common and best developed in the plains (Chicarro et al. 1985), which suggests that the mechanical properties of this unit facilitated ridge nucleation.

In many areas, the ridges form a subparallel or concentric pattern with a regular spacing (Wise et al. 1979; Saunders and Gregory 1980; Saunders et al. 1981; Maxwell 1982; Watters 1991) that provides a constraint on relationships between the competence and thickness of the volcanic materials and the structure of the underlying lithosphere (Saunders and Gregory 1980; Saunders et al. 1980, 1981; Watters 1991; Zuber and Aist 1990). For current estimates of ridged plains thickness, strength contrasts between the plains and underlying megaregolith from 1 to 3 orders of magnitude are allowable, which does not meaningfully limit the range of possible rheologies of these units. Improved observational estimates of flow thicknesses would further constrain both the mechanical properties of the plains at the time of ridge formation and the volume of plains-forming volcanism.

VII. FUTURE STUDIES

As the Mars community awaits new data from the Mars Observer mission, it is useful to consider what issues in Martian volcanology have remained unresolved with Viking and Earth-based data. A selection of appropriate questions that we hope will occupy the next stage in our attempts to understand volcanic processes on Mars include the following.

1. What was the range of magma chemistries that produced the volcanic landforms that are now exposed at the surface? In particular, the discovery of silicic materials on Mars would not only have significant impact on our understanding of magmatic evolution on Mars, but would also greatly increase the expected diversity of explosive volcanic deposits that should exist in Mars (Francis and Wood 1982). Conversely, if the leading candidate deposits that could be silicic (such as the Amazonis deposits described by Scott and Tanaka [1982]) are found to be basaltic, then further considerations of the physics of explosive basaltic activity must be pursued.

2. What are the absolute ages of volcanic rocks associated with the major eruptive episodes on Mars? Radiometric ages would establish the amount, frequency and consequence of the release of juvenile volatiles into the evolving Martian atmosphere. A better understanding of the frequency of volcanic activity on Mars, and the probable time intervals between successive eruptions of individual volcanoes (yr or Myr?) would provide climate models with important constraints on the evolution of the Martian atmosphere through geologic time.

3. What are the implications of volcano growth with time for the local-stress history of the lithosphere in the vicinity of the major volcanic constructs, and how did this changing load affect the internal structure of each volcano? Loading of the volcanic pile may well have altered the internal plumbing system, promoting a redistribution of dikes, vents and fissures, as well as a change in the geometry and location of the magma chamber.

4. What was the origin of the Tharsis and Elysium Provinces? Understanding the origin of these young volcanic centers may provide an explanation as to why constructive volcanic activity evidently terminated at an earlier time in the Martian southern hemisphere than in the north, and why only pyroclastic volcanoes appear to have formed in this region, as opposed to the probable polygenic activity in the northern plains.

5. What information can the distribution of tectonic features within summit calderas provide about caldera formation and the evolution of Martian magma chambers?
It is likely that the current debates concerning the origin of the Olympus Mons escarpment, and the probable volcanic origin of the ridged-plains materials, will also continue. Several of the above specific questions will probably only be answered with the return to Earth of carefully selected fresh Martian samples. We nevertheless eagerly await the thermal infrared, high-resolution camera, laser altimetry, gravity and gamma-ray data from Mars Observer, and also feel that the careful use of Earth-based radar and spectral data, along with the continued analysis of the Viking image data set, will provide additional insights into the physical volcanology of Mars.

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